Numerical simulation of the development of tropical cyclones with a ten-level model. Part II

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ABSTRACT

The importance of some different physical processes and parameters, pertinent to the development of tropical storms, is studied with the aid of the numerical model presented in an earlier paper (Sundqvist, 1970). The developing tropical cyclone exhibited and discussed in detail in the previous paper is used as a reference case for the present experiments.

The results show that the rate of intensification becomes significantly higher when radiational cooling is taken into account. On the other hand, the peak intensity appears to be unaffected.

In another experiment, where a hypothetical poleward movement of the vortex is introduced, it is demonstrated that a mature hurricane may remain quite intense to rather high latitudes, provided other conditions are unchanged.

The decisive role of the sea surface temperature is clearly exhibited in three other applications. For one and the same vertical stratification of the tropical atmosphere the maximum swirling velocity never exceeds 25 m/s when the water temperature is 26°C, while a full-fledged hurricane develops at 27.5°C.

In an attempt to reduce the highest wind speeds of a mature storm by enhancing the heating artificially (cloud seeding), the results indicate that the approach yields the desired tendency in the evolution. However, no changes of significance are observed in the maximum wind during the hypothetical operation that was assumed to last for 40 hours.

1. Introduction

In a recent paper (Sundqvist, 1970; henceforth referred to as P1) the author described an axi-symmetric balanced model designed for numerical simulation of the development of tropical cyclones. The results of one experiment which were presented and discussed in detail in P1 showed that the model is capable of simulating a great number of the characteristic features of real tropical cyclones. The examination also revealed a pronounced internal consistency of the behaviour of the model vortex during its evolution. It is therefore concluded that this model may be used with confidence for comparative experiments, although the horizontal resolution is relatively crude ($\Delta r = 25$ km). The purpose of the present paper is to carry through some similar applications, whereby we regard the experiment presented in P1 as a control or reference case.

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We may divide the experiments into three classes:

(i) experiments with additional or changed physical processes in the model;
(ii) experiments for studying the influence of various external parameters on the evolution;
(iii) experiments aimed for investigating possibilities to alter the evolution artificially.

A few comparisons of each category will be considered. From class (i) an experiment with radiational cooling will be studied. Under class (ii) we shall investigate the effect on the evolution when the vortex is subject to a poleward motion and furthermore we shall study how the sea surface temperature influences the development. In class (iii) an experiment concerning artificially augmented convection will be performed. In addition, a comparison—class (i)—of the two versions M1 and M2 of the condensation model was previously discussed in P1 (subsection 6.2).

We shall be satisfied in merely presenting a few features that most clearly exhibit the deviation from the reference case (henceforth denoted case A). In fact it has turned out that the intensity in terms of the maximum tangential wind (or central surface pressure) in a majority of the cases is the most useful quantity for this purpose. Generally it is difficult to find differences in the detailed structure of the vortex or its energy budget that make conclusive deductions possible.

As this paper is considered a continuation of P1, a description of the model will not be given here. It is deemed sufficient only to repeat some of the parameters of the case presented in detail in P1. In case A we thus have

$f = 5 \cdot 10^{-5}$ sec^{-1};	the Coriolis parameter at 20° N;
$T_w = 27.5$°C;	the temperature of the sea surface;
$q_w(r_{max}) = 23.6 \cdot 10^{-3}$;	the specific humidity of the sea surface at $r = r_{max}$;
$T_s(r_{max}) = 27.5$°C;	the surface air temperature at $r = r_{max}$.

Both f and T_w are constant along the radius, while q_w, the saturation humidity with respect to T_w, is inversely proportional to the surface pressure according to the definition of specific humidity. Furthermore the initial vortex is barotropic and has a maximum wind of 15 m/s at radius 200 km.

2. An experiment including cooling due to radiation

The rate of cooling due to radiation was briefly discussed in P1 (section 3). It was inferred that, although this cooling may not contribute significantly to the heat budget in the model vortex, ($r_{max} = 600$ km), its presence in the cloud-free regions might nevertheless increase the differential heating in the system, hence resulting in a more rapid intensification.

Clear air infrared cooling rates have been computed according to Rodgers & Walshaw (1966) for a few different vertical profiles of temperature and humidity, taken from case A. A mean vertical distribution of the cooling

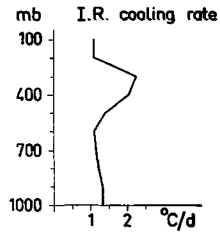

Fig. 1. Vertical profile of the infrared cooling rate (in °C/day) taken into account at all points outside the region of condensation.

rate, Fig. 1, was constructed from the profiles obtained, which in fact did not deviate very much from each other. This cooling is introduced at each point outside the region of condensation. Otherwise the model is the same as for case A.

During the development, the magnitude of the total cooling as integrated over the domain used in the computation is about 10–15% of the total heating by condensation in the system; measured per unit mass, the ratio is about 1:100. In spite of the relatively small contribution by the cooling, its enhancement of the differential heating causes a noticeable increase in the rate of intensification. This is shown in Figs. 2 and 3 which portray the 900 mb maximum tangential wind and the total (0–600 km) kinetic energy versus time respectively. We notice that this storm matures earlier than the one in case A so that there is no difference in the peak values of the maximum wind in the two cases. On the other hand the increased differential heating has made the conversion of potential to kinetic energy easier. The difference between the maximum and original amounts of total kinetic energy of the system

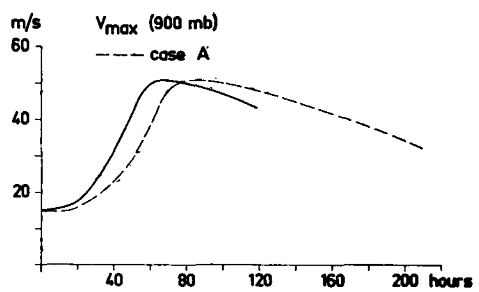

Fig. 2. The 900 mb maximum tangential wind versus time for the case with radiational cooling included (——) and for case A (– – –).

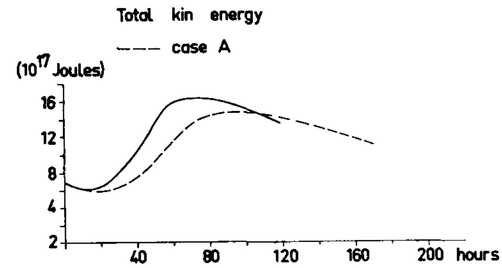

Fig. 3. Total kinetic energy (0–600 km) versus time for the same cases as in Fig. 2. The unit is 10^{17} Joules.

is about 25% higher in this case than in case A. In their study of the generation of available potential energy in hurricane Hilda, Anthes & Johnson (1968) deduced that the infrared cooling contributed to the generation by about 25%. Since our figure above gives the *net* increase of kinetic energy, it seems somewhat high compared with the value given by Anthes & Johnson (1968). They assumed, however, that the cooling essentially took place outside 500 km, while in the present experiment the cooling starts already at 200–250 km radius. Thus the overall differential heating induced by the cooling becomes more pronounced in the present case. As no cirrus shield, which reduces the infrared cooling, is assumed to exist outside the region of precipitation, we may infer that the present experiment exhibits the upper limit of the effects to be expected by radiational cooling.

3. Effects of changes in external parameters

3.1. *Poleward motion of the mature cyclone*

Before we study the effect of changes of the sea surface temperature that may arise as the cyclone moves poleward for instance, it is of interest to examine how strongly a steadily increasing Coriolis parameter affects the evolution of the storm.

As we are dealing with an axi-symmetric system within which f has thus to be constant, we must simplify the simulated poleward motion by assuming that the entire vortex experiences the same change in f.

In the present experiment the Coriolis parameter is given a change at each time step corresponding to a northward movement with a speed of 20 km/h ($\simeq 4.3$ °lat/day). The state at 84^h of case A is chosen as the initial one; we are thus considering the poleward motion of a mature hurricane. The integration is carried out over 75 hours (from 20° N to 33.5° N) and the resulting v_{max} variation is shown in Fig. 4 together with case A. It is seen that the intensity (in terms of max. wind) declines more rapidly in the present case than in case A, the difference being fairly small, however. This indicates that a hurricane would remain quite intense during a poleward motion even over large distances, provided other conditions are unchanged.

The faster decay obtained in the present experiment is due to a reduced frictional inflow that is a consequence of the increasing Coriolis parameter. The region where small values of the absolute vorticity appear does not look significantly different from the one in case A (subsection 6.3. in P1). It is therefore inferred that the accelerated decline in intensity due to inertial stabilization comes entirely via a diminished frictional inflow.

3.2. *Experiments with changes in sea surface temperature*

Palmén (1948) showed in his study on hurricane formation that a sea surface temperature of 26–27°C is a necessary condition for the development of hurricanes. Several other investigations (e.g. Fischer, 1958; Miller, 1958; Perlroth, 1962, 1967, 1969; Leipper, 1967) clearly indicate that the intensity of tropical cyclones is closely related to the temperature of the sea surface. Ooyama (1969) performed a series of experiments with his numerical

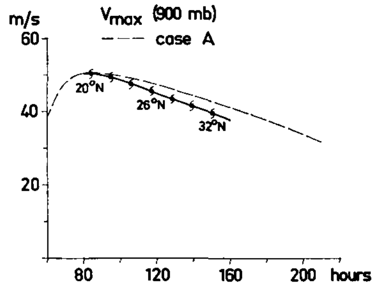

Fig. 4. The 900 mb maximum tangential wind versus time when f varies corresponding to a northward movement of the storm with a speed of 20 km/h (——); the latitudinal positions are indicated for every second degree. ---, Case A.

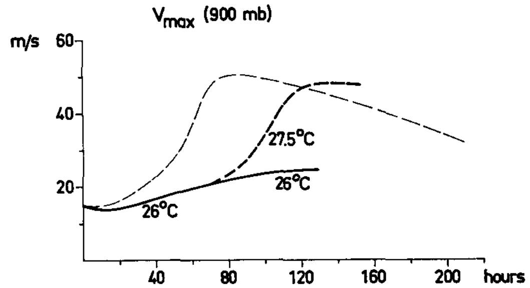

Fig. 5. The variation of v_{max} with time; (*a*) in case A where $T_w = 27.5°C$ (---); (*b*) for $T_w = 26°C$ (——); (*c*) for $T_w = 27.5°C$ from 71^h (- - -). The integration (*c*) is started at 68^h on (*b*) and increases with 0.5°C/h between 68^h and 71^h.

model in order to study the effects of the sea surface temperature. The results of those experiments also demonstrated that the behaviour of the vortex markedly depends on T_w.

In the first application with the present model in this context we put $T_w = 26°C$ and also $T_s(r_{max}) = 26°C$. The time variation of v_{max} is shown by the full line in Fig. 5. In good accordance with Palmén's conclusion, the hurricane development is inhibited under these circumstances; the maximum wind never reaches 25 m/s.

It is also interesting to study how the radii of v_{max} and w_{max} respectively—$r(v_{max})$, $r(w_{max})$—change in this case. Fig. 6 shows that $r(v_{max})$ decreases to practically the same value as in case A and that $r(w_{max})$ is smaller than in case A. The latter effect is due to the slower rotation of this case; the frictional inflow of the quasi-balanced model is inversely proportional to the absolute vorticity and since the relative vorticity is strongest in the central parts of the vortex it follows that the maximum inflow moves closer to the centre as the circulation weakens. As the comparison displayed in Fig. 6 indicates that the intensity influences $r(v_{max})$ very little, it seems reasonable to set the radius of the maximum wind of the order 50 km already in the initial vortex.

In the second experiment in this series we take (arbitrarily) the 68^h state of the 26-degree case as the initial state for an integration where we let T_w and $T_s(r_{max})$ increase to 27.5°C at a rate of 0.5°C/h. Thus from 71^h the conditions are exactly the same as in case A. The resulting development, depicted by the thick dashed line in Fig. 6, then becomes practically the same as the one of case A. Nearly the same peak intensity is also attained (48.5 m/s).

It is usually observed that disturbances move westward over large distances of the eastern halves of the oceans without significant changes in intensity (see, e.g., Carlson, 1969) i.e., the part where the sea surface temperature generally is relatively low and has small longitudinal variations. Those disturbances that finally become hurricanes then undergo a rapid deepening to a mature stage. The combined results of the two experiments above thus agree very well with such characteristics of real hurricane development.

In the third application, the response to fluctuations in T_w are studied. In this experiment we start at 84^h of case A and let T_w and $T_s(r_{max})$ decrease with 0.5°C/h until 26°C is reached (at 87^h). Then this temperature of the sea surface is kept constant for 12 hours. At 99^h a temperature rise, like the earlier decrease, starts and from 102^h the conditions are thus again the same as in case A. The variation of v_{max} with time is shown in Fig. 7. The maximum wind of the model storm decreases with about 13 m/s during the 12 hour period. The corresponding rise in the central surface pressure is about 22 mb (from 951 to 973 mb). The results indicate that the response is more immediate to the drop than it is to the rise in temperature; it takes somewhat more than 30 hours to restore a new mature stage after the environmental conditions have become the same as before 84^h.

The response to the fluctuation in T_w is even more conspicuous in the rate of precipitation. Fig 8 shows that the rate of mean precipitation inside radius 100 km is reduced by about 60%. It is furthermore interesting to note that the response is much less pronounced farther out (100–200 km). This fact indicates that a

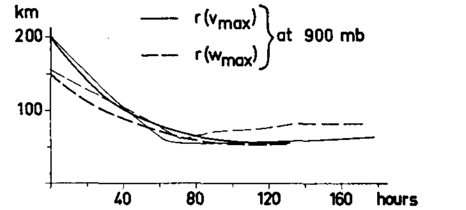

Fig. 6. The time variation of the radius of v_{max} (——) and radius of w_{max} (- - -) for the case T_w = 26°C. The corresponding variations in case A ($T_w = 27.5°C$) are shown by thin lines.

drop in T_w only beneath the core region of the storm (inside radius 100 km or 200 km, say) may bring about almost as substantial an intensity decrease as in the present experiment. This deduction also agrees with the results of Ooyama's (1969) experiments which showed that the development by and large remained the same, with $T_w = 27.5°C$ unchanged within 150 km or 300 km radius, despite significantly different values of T_w were applied outside those radii. Consequently even relatively small scale variations in sea surface temperature may be reflected in the intensity of tropical cyclones.

The above results show that frequent, accurate observations of the surface temperature of the tropical oceans would be of prominent value; not only for providing possibilities to verify results of numerical models, but also, it seems, for the possibility of indicating expected intensity variations in actual forecasts of a storm's movement. At present we know nothing about the feedback of a changing sea surface temperature on the movement of tropical storms. On the other hand, as demonstrated, the storm itself is sensitive to such changes regardless of its path.

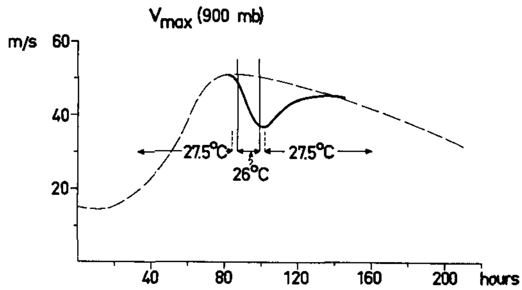

Fig. 7. Maximum tangential wind versus time (——) when T_w varies. T_w starts to decrease and to increase respectively at 84^h of case A and at 99^h in the same way as described in Fig. 5. ---, case A.

4. An attempt to modify the storm artificially

Although a number of simplifications have been made in deducing the present numerical model on hurricane development, this may nevertheless be useful for rather direct practical applications like investigations of the possibility of modifying real tropical storms by artificial means. Applications of this kind will provide a guidance for when and where to perform such operations in order to attain the best possible effects. In addition it is of great practical importance to find out for how long a time a storm has to be treated artificially, before it possibly continues a desired evolution by itself.

At present, cloud seeding seems to be the most realistic means for the purpose of exercising some control on the behaviour of tropical storms. The results of field experiments with seeding of individual cumuli indicate that seeded clouds precipitate significantly more than do undisturbed clouds (Simpson & Woodley, 1969). Therefore it is interesting to see how the model storm reacts to a hypothetical seeding.

In P1 (subsection 2.2) it was inferred that the heating should be augmented far out from the centre in order to cause a reduction in intensity of the storm. As the seeding method requires that clouds already exist, the fringe of the convective region should be the best place for this operation.

In the specific experiment an additional heating is introduced at each time step in the two outermost grid points of the convective region. If the heating rate as computed from the condensation model (section 3 in P1) corresponds to a precipitation rate less than 2 mm/h then this heating is increased by a factor of 4 at each level. If the precipitation lies between 2 and 3 mm/h the heating is increased by a factor of 2. For rates higher than 3 mm/h the heating is not changed. The integration is started from 84^h of case A and the addition of

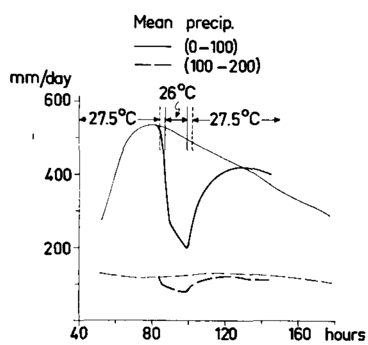

Fig. 8. The rate of mean precipitation versus time (in mm/day) in the intervals 0–100 km (——) and 100–200 km ---). The thin lines show the corresponding variations in case A.

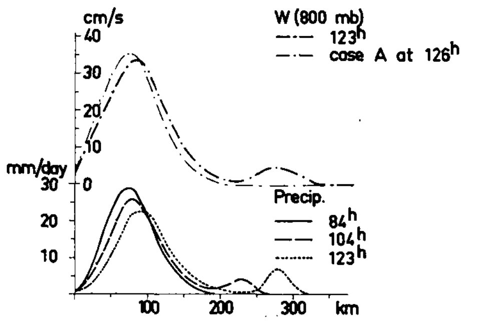

Fig. 9. Display of effects caused by artificially augmented heating in the outer part of the convective region, starting from 84^h of case A. Upper part shows the 800 mb vertical velocity versus radius at 123^h (–•–•–) and that of case A at 126^h –·–·–). Lower part shows the radial distribution of the rate of precipitation (in mm/h) at 84^h (——), 104^h (– – –) and 123^h (• • • •).

heat is made throughout the period of integration which ended at 123^h. Both with regard to the magnitude of the heat increase and to the duration of the operation this is probably more than is practically feasible. Yet there is no noticeable reduction of the maximum wind in comparison with case A.

A closer examination of the results of this experiment reveals, however, that the manipulation has produced the right tendency for the wanted effects to occur. The radial distribution of the precipitation intensity of the present case is shown for three different times (84^h, 96^h and 123^h) in Fig. 9. It is clearly seen how the enhanced heat source—around radius 200 km in the beginning—makes the convective area expand gradually. At 126^h of case A, the release of latent heat is still confined to the area inside radius 200 km. Furthermore it is seen that the secondary maximum of heating causes a reduction and outward migration of the primary maximum. This fact also appears in comparison with case A. In the upper part of Fig. 9 the radial distribution of the 800 mb vertical velocity at 123^h and that of case A at 126^h are displayed. The pattern of the meridional circulation in this case is worth noticing. As indicated by the w-curve in Fig. 9, the low level radial inflow has increased relatively much outside 300 km resulting in a sedondary maximum in the vertical velocity between 250 and 300 km. Thus the artificially increased heating rate at a large distance from the centre tends to cause a cut off and eventually a subsequent choking of the core region.

In view of the above assumption that the enhancement of the convection (and the duration of the impulse) in the present experiment is at least of the order of what can be induced in real situations, the results thus imply that it would be difficult to bring about a significant decrease in intensity of a *mature* hurricane by means of cloud seeding. In a mature tropical cyclone the release of latent heat in the inner region is so enormous that the artificial increase almost becomes negligible; especially as this increase of the convection also has a relatively small radial extent. (The experiment presented in section 2 showed that the cooling, although very small per unit mass, affected the development appreciably due to its vast appearance.)

It would be interesting to repeat the above experiment, starting however in the middle of the deepening stage of the case A storm. Then the heating rate in the core region is only half that of the mature stage (see, e.g., Fig. 18 in P1). The effects of the artificial heat source would thus, by and large, become doubled compared with the experiment above.

Acknowledgement

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REFERENCES

Anthes, R. A. & Johnson, D. R. 1968. The generation of available potential energy in hurricane Hilda (1964). *Mon. Wea. Rev. 96*, 291–302.

Carlson, T. N. 1969. Some remarks on African disturbances and their progress over the tropical Atlantic. *Mon. Wea. Rev. 97*, 716–726.

Fisher, E. L. 1958. Hurricanes and the sea surface temperature field. *J. Meteor. 15*, 328–333.

Leipper, D. G. 1967. Observed ocean conditions and hurricane Hilda, 1964. *J. Atm. Sci. 24*, 182–196.

Miller, B. I. 1958. On the maximum intensity of hurricanes. *J. Meteor. 15*, 184–195.

Ooyama, K. 1969. Numerical simulation of the life cycle of tropical cyclones. *J. Atm. Sci.* 26, 3–40. (First published in *Geophysical Sci. Lab. Report No. TR 67–9*, 1967, New York University.)
Palmén, E. 1948. On the formation and structure of tropical hurricanes. *Geophysica 3*, 26–38.
Perlroth, I. 1962. Relationship of central pressure of hurricane Esther (1961) and the sea surface temperature field. *Tellus 14*, 403–408.
— 1967. Hurricane behaviour as related to oceanographic environmental conditions. *Tellus 19*, 258–267.
— 1969. Effects of oceanographic media on equatorial Atlantic hurricanes. *Tellus 21*, 230–244.
Rodgers, C. D. & Walshaw, C. D. 1966. The computation of infra-red cooling rate in planetary atmospheres. *Quart. J. R. Met. Soc. 92*, 67–92.
Simpson, J. & Woodley, W. L. 1969. Modification experiments on tropical cumulus clouds (Abstract of paper presented at the Sixth Technical Conference on Hurricanes, Dec. 2–4, 1969, Miami, Fla) *Bull. Am. Meteor. Soc. 50*, 779–780.
Sundqvist, H. 1970. Numerical simulation of the development of tropical cyclones with a ten-level model. Part I. *Tellus 22*.

ЧИСЛЕННОЕ МОДЕЛИРОВАНИЕ ЭВОЛЮЦИИ ТРОПИЧЕСКИХ ЦИКЛОНОВ С ПОМОЩЬЮ ДЕСЯТИУРОВЕННОЙ МОДЕЛИ: ЧАСТЬ II.

С помощью численной модели, описанной в предыдущей работе (Sundqvist, 1970), изучается значение различных физических процессов и параметров, связанных с эволюцией тропических циклонов. Развитие тропического циклона, которое демонстрировалось и детально обсуждалось в предыдущей статье, используется для сравнения с настоящими экспериментами. Результаты показывают, что скорость интенсификации становится значительно больше при учете радиационного выхолаживания. С другой стороны, максимальная интенсивность при этом, по-видимому, не меняется. В другом эксперименте, где вводится гипотетическое движение вихря к полюсу, показано, что тайфун может оставаться довольно интенсивным до весьма высоких широт, если остальные условия не меняются. Решающая роль температуры поверхности океана явно обнаруживается при трех других экспериментах. Для одной и той же вертикальной стратификации тропической атмосферы максимум тангенциальной скорости никогда не превосходит 25 *м*/сек при температуре воды 26°C, тогда как ярко выраженный тайфун развивается при 27,5°C. Попытка уменьшить максимальные скорости ветра тропического циклона путем искусственного увеличения нагревания (с помощью воздействия на облака) показала возможность такой тенденции. Однако никаких значительных изменений максимальной скорости ветра во время численного эксперимента, длившегося 40 часов, не наблюдалось.